Trimingham: structural architecture of the Cromer Ridge Push Moraine complex and controls for landslide geohazards

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1. Introduction

One of the most striking geomorphological features of the north Norfolk area is the Cromer Ridge. It extends eastwards from Thursford, near Fakenham, to Trimingham where it intersects the North Sea coast. The ridge has been interpreted in-part, as a push moraine formed at the southern margins of the Middle Pleistocene ice sheet and is draped on its ice-distal southern flank by extensive spreads of outwash sand and gravel (Hart, 1990; Pawley *et al*., 2005). Coastal sections between Trimingham and Overstrand offer a rare opportunity to examine the internal architecture of this structure. The focus of this chapter is to document the structural geology of this landform, and examine the influence of its structure on coastal stability and landslide geohazards.

Section Location: park on the western outskirts of Trimingham in the lay-by (TG 275 388) and head northeast along footpath to cliff edge. Descend cliff to beach level. Sections are mainly situated to the west. Site not accessible at high tide. Extreme care should be taken when descending the cliffs.

2. Geological background

The stratigraphy of Trimingham cliffs was originally surveyed by Reid (1882), but has presented something of a geological puzzle to many scientists owing to the extended layer-cake succession (Lee, 2003), the intense glacitectonic deformation (Hart, 1990), and the frequent obscuring of exposures by large rotational landslides (Hutchinson, 1976). Consequently, several different stratigraphic models have been presented (Hart, 1992; Lunkka, 1994) and the true stratigraphic significance of the succession is somewhat confused (Roberts and Hart, 2000). Due to limited exposure, unravelling the stratigraphic complexity is key to resolving the major structural features and the geometric relationship between the major units.

In this study, coastal sections were examined over a ten year period between 1996 and 2006 to take advantage of cliff falls and coastal erosion which created new, often temporary, exposures. Individual sections were logged paying attention to lithology, unit geometry and tectonic and sedimentary structure. Bulk samples were also collected with particle size analysis and heavy mineral analysis performed systematically to assist with correlating units and to create a local stratigraphic model. Comparison with the regional stratigraphic succession enabled the deformation to be unravelled and the reconstruction of the pre-deformation stratigraphy (Table 13.1). Four distinctive till units were recognised (cf. Hart, 1992; Lunkka, 1994), specifically the Happisburgh Till, Walcott Till, Bacton Green Till and Weybourne Town Till, and these have been recognised more widely where they form distinctive and mappable geological units (Lee *et al*., 2004; Hamblin *et al*., 2005). Each of the tills is separated by glaciolacustrine and/or fluvio-deltaic deposits of variable thickness and the sedimentology conforms generally to observations of lithofacies and unit thicknesses noted elsewhere in northeast Norfolk (Lee, 2003; Lee *et al*., 2008a; Lunkka, 1994). The exception to this is the much thicker glaciolacustrine succession demonstrated by Hart (1992) that occurs between the Walcott and Bacton Green tills (Lee, 2003). This succession includes rhythmically bedded and sometimes varved silts and clays, and beds of sand and marl (Hart, 1992; Lee, 2003).

3. Structure of the Trimingham-Overstrand section

The Trimingham to Overstrand coastal section can be sub-divided into three structural domains based upon the style of deformation and their cross-cutting relationships (Figure 13.1). Domain 3, located at the northern end of the coastal traverse, encompasses the large glacitectonic rafts at Overstrand. They are discussed elsewhere within this field guide (see Chapter 12).

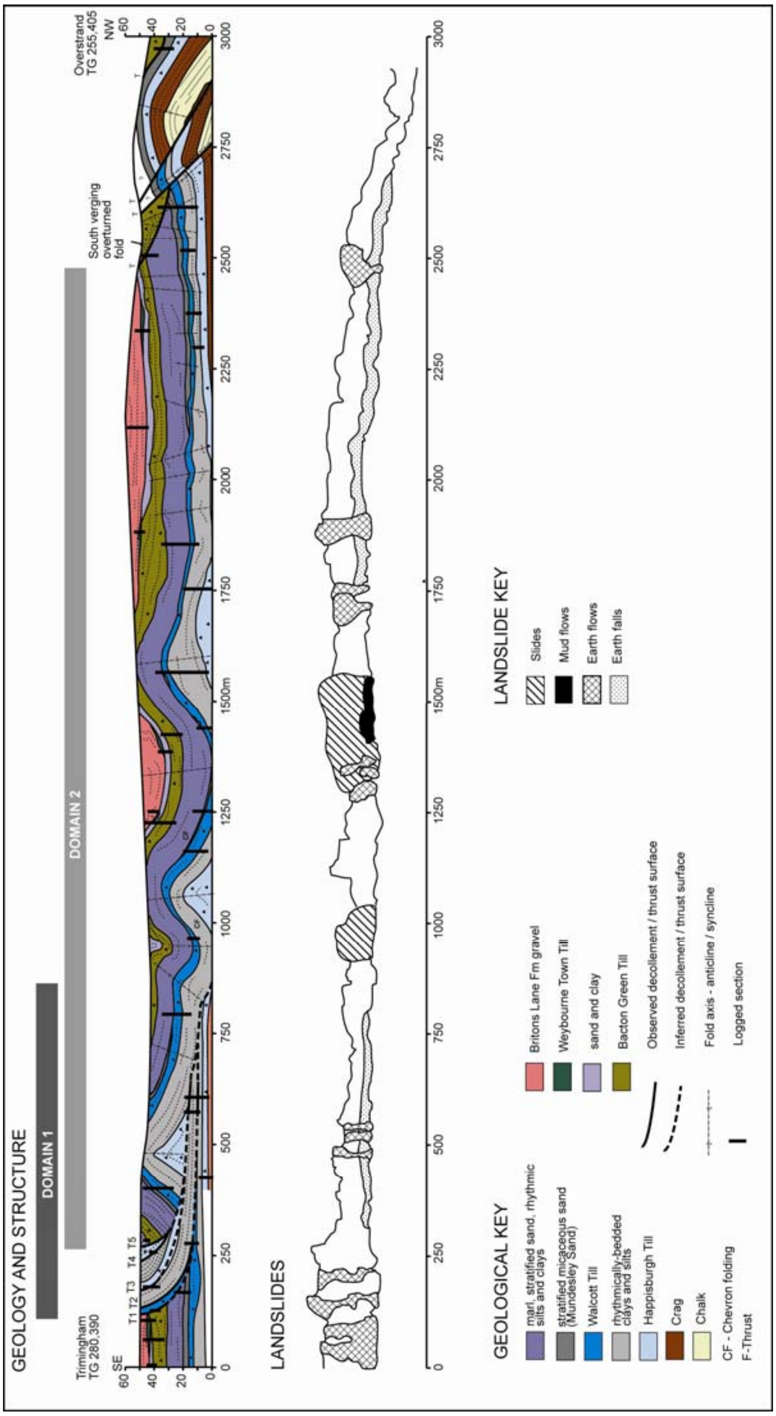

Figure 13.1. Cross-sections of the Trimingham to Overstrand coastal sections showing the interpreted geology and landslides (modified from Hart, 1990)

3.1 Description of Structural Domains

Structural Domain 1 (SD1) occurs between 825-100 m at the southeastern end of the coastal section and cross-cuts undeformed glacial and preglacial deposits that occur beneath the village of Trimingham. The domain is bounded by a basal décollement surface (T1) separating it from undeformed glacial and preglacial deposits, and an upper glide horizon composed of a series of anastomosing fault surfaces (parts of T3 and T5) that branch-out progressively from T1. The structure of the domain is characterised by a series of sub-horizontal décollement surfaces (T1-T5) that ramp upwards through the sediment pile in a southerly direction taking the form of a series of listric thrust and high-angle reverse faults, forming an imbricate hinterland-dipping duplex. A characteristic feature of this duplex is that it resulted in the thrust stacking of older parts of the pre-glacial and glacial succession onto much younger parts of the glacigenic sequence. T1 forms the basal detachment surface and extends through a sequence of varved glaciolacustrine silts and clays that crop-out between the Happisburgh and Walcott tills, before propagating upwards through the sediment pile along a series of tectonic 'ramps' and 'flats'. Field observations reveal that T1 extends laterally southwards along a series of bedding planes ('flats') between clay and silt couplets until it reaches a thick less competent bed and then 'ramps' upwards through the varved sequence. When the detachment reaches a thicker and more competent silt bed the décollement surface continues to extend laterally along the 'flat', thus enabling the detachment to rise gradually through the sediment pile. Walcott Till is preserved within the hanging-wall of T1 between 250-100 m.

T2 forms an anastomosing detachment surface that branches from T1 at 250m with the northern part of T1 décollement thus becoming part of T2. The hanging-wall block of T2 is composed entirely of varved clays which over-thrust Walcott Till between 420-180 m placing older lithologies over younger lithologies. The basal surface of the hanging-wall exhibits two distinctive sets of cross-cutting slickenslides. Kinematically they demonstrate a subtle shift in the direction of translocation along the décollement surface: firstly, from north to south, and then secondly, northwest to southeast (Figure 13.2).

Décollement surface T3 forms an additional major detachment surface and can be traced discontinuously between 825-190 m. The hanging-wall block is composed of Happisburgh Till and varved clays which have been thrust over the younger deposits within the hanging-wall block of T2. The inclusion of Happisburgh Till shows that the active décollement surface extended both northwards and downwards through the sediment profile during this later stage entraining progressively older lithologies into the duplex. T4 is a small high-angle detachment that was recorded at 270-250 m. Contained within the hanging-wall of the reverse fault is Happisburgh Till which may have been detached from between 400-350 m. T5 forms the upper glide horizon to the duplex structure with the hanging-wall composed of Domain 2.

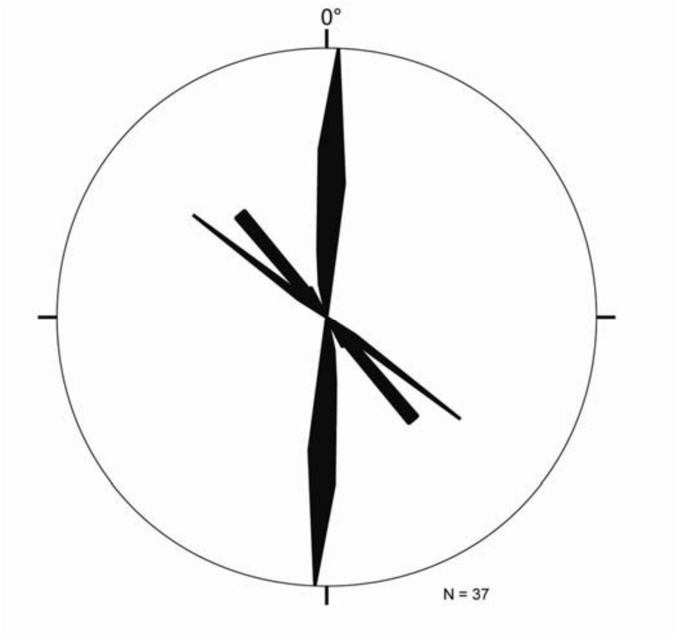

Figure 13.2. Rose diagram showing the trend of slickensides along shear plane T2

Domain 2 occurs between 2650-270 m and forms the hanging-wall block of T5 which represents the upper glide horizon to the imbricate duplex described within Domain 1. Domain 2 is characterised by a series of large-scale, open, symmetrical to slightly asymmetrical, synclines and anticlines that gently deform the entire glacial sediment pile beneath the Britons Lane Formation which forms the cap to the glacigenic succession. The total amount of shortening within Domain 2 which has been accommodated by this folding was calculated using the Heime 'unrolling' Method (Figure 13.3) and is approximately 16.8%. This is considerably less shortening than that estimated by Hart (1990) which was 32%.

At the northwestern end of the section between 2650-1750 m, folding is typically open, with wavelengths varying between 20-80 m and amplitudes of up-to 10m. Beyond 1750 m, the wavelength (up-to 250-500 m) and amplitude (up-to 20m) of the folds are more variable but generally increase in scale, forming several large and distinctive synclinal basins. Axial surfaces are upright to moderately inclined, plunging towards either the south to southwest or north to northeast. The strike of the axial surfaces of the folds are aligned broadly west-northwest to east-southeast. Adjacent to the largest syncline between 1600-1100 m, smaller-scale shortening has been accommodated within beds of rhythmically-bedded or laminated silts and clays leading to the development of distinctive chevron folds (e.g. 1500 m, 1280 m, 1170 m). At the southern end of Domain 2, beds have been deformed into a large south-verging asymmetrical anticline.

$$e = \frac{A'B' - AB}{AB}.100$$

$$e = \frac{2410 - 2900}{2900}.100$$

e = 16.8% shortening

Figure 13.3. The Heime 'unrolling' Method for calculating the degree of shortening within a folded sequence. The technique was employed on the distinctive 'lacustrine marl bed' that occurs between the Walcott and Bacton Green tills. *e* is the amount of shortening, *A'B'* corresponds to the deformed length and *AB* the original undeformed length of the sequence.

4. Landslide Geomorphology

4.1 Description of landslides

Coastal landslides (Figure 13.4) form an active geomorphological component along much of the North Norfolk coast, but the Trimingham area in particular is highly prone to instability (Hobbs *et al*., 2008; Hutchinson, 1976). The distribution and type of landslides are controlled in-part by the level of coastal management (e.g. coastal defences) plus the lithology and structure of deposits within the cliffs (Figure 13.1).

Between 0-270 m, landslides consist of a number of individual and coalescent earth flows and falls in the upper cliff, that cascade down the cliff profile and are deposited in low gradient or flat areas. The first 250m, includes cliff protected by coast-parallel wooden revetments. The dominant landslide style is earth flows. Flows create a distinctive cliff profile with scallop-shaped bowls created within the upper erosional zone of the landslide. Repeated failure and flow causes these bowls to coalesce and the upper cliff to recede. This creates a distinctive bench feature midway down the cliff. Flows coincide with the near vertical propagation of the basal décollement surfaces (T1-F3) towards the top of the cliff section (i.e. SD1). Bedding within this zone is sub-vertical with groundwater seepage along bedding and structural surfaces causing materials to liquefy and flow. Continuing west towards the western end of SD1, falls are the dominant landslide mechanism with small scree slopes and debris cones armouring the base of the cliff. Falls appear to be particularly common in association with thrust slices of laminated/varved sediment where bedding surfaces now dips seaward.

Landsliding along the remainder of the coastal traverse is largely controlled by the open synclines and anticlines that characterise SD2. It consists mainly of large slides, with localised debris flows and falls. Large composite slides occur between 900-1025 m and 1300-1570 m, and are primarily deep-seated movements that have a dominant rotational component, but are in-part translational. They coincide with the large open synclines and are constrained laterally by structural anticlines that form sharp morphological buttresses. Between 1300-1570 m, the basal shear surface of the slide extends beneath the beach platform following the form of the syncline with an up-thrusting component of movement at the toe. Repeated sliding has formed a deeply-incised to elongate embayment that is arcuate in plan-form, with backscarps aligned parallel to the coastline. The deep-seated slide has resulted in back-tilt and extension features on the cliff slope, which have led to the formation of seasonal ponds. Seepage from these ponds often leads to the reactivation of shear planes and can cause minor sliding, or the development of minor earth and mud flows. This is especially the case where seepage occurs within the stratified sand and clays that form the Sheringham Cliffs Formation. Further westwards, landsliding is characterised predominantly by earth flows and falls.

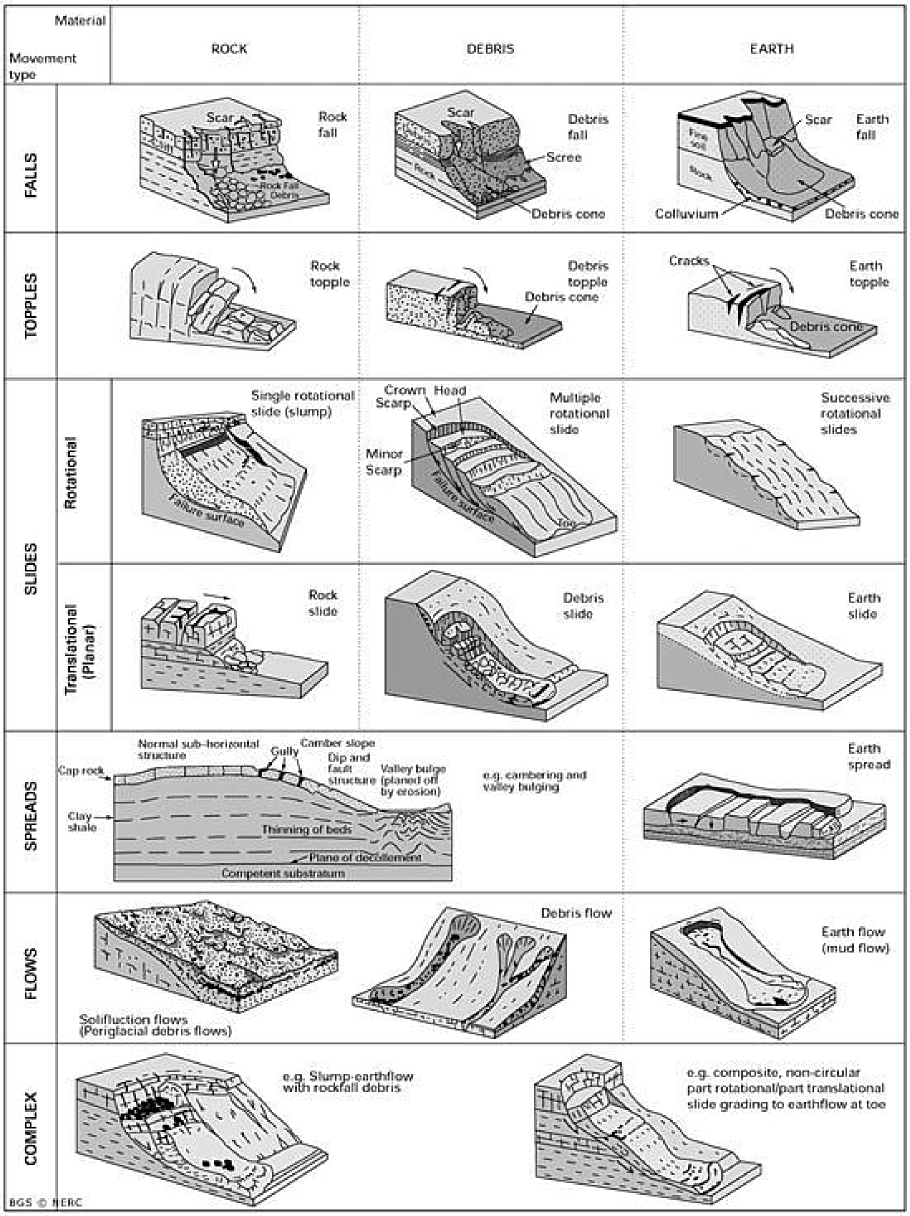

Figure 13.4. Categories of coastal landslide

Earth flows are localised and coincide with smaller open syncline structures (e.g. 1750 m, 1900 m, 2500 m). Earth falls by contrast occur along much of the mid-high parts of the cliff-line where they occur within very gently folded strata. Failure occurs typically by undermining of more competent till and clay beds by seepage from a confined aquifer within sandy units – in this case the sands within the Sheringham Cliffs Formation, which occur at several different stratigraphic levels. The more competent deposits within the lower part of the cliff (e.g. Happisburgh and Walcott tills and intervening clays) are less prone to this type of failure and form steep cliff faces. Scree developed from earth falls typically cascades to the base of the cliff, or is deposited temporarily on flat bench areas in the cliffs profile that form in competent lithologies.

5. Conclusions and points for discussion

- The internal structural architecture of the Cromer Ridge between Trimingham and Overstrand is typical of many push moraines (e.g. van der Wateren, 1995) in that it has been largely developed by piggyback thrusting. In this example, several thrust slices have been thrust over one another to form a stacked duplex structure as an earlier study demonstrated (Hart, 1990). It contains two structural domains: (SD1) characterised by the large-scale décollement surfaces that develop into listric thrust faults within the duplex; (SD2) characterised by large-scale open folding.

- Landslide development along the coastal traverse is strongly controlled by glacitectonic structure. Thrusts (i.e. synclines) offer natural failure planes for landsliding as well as acting as planes that enable groundwater to infiltrate and migrate around the sequence. Thrusting also alters the geometry of discontinuities such as bedding and jointing - at Trimingham, sub-vertically aligned beds of diamicton are highly prone to failure, generating earth flows and falls. The movement of sediment blocks of different permeability relative to each other during thrusting (and landsliding) leads to the development of highly localised and confined groundwater conditions which often accentuates landslide hazards. Synclines within the sequence also offer topographic lows within which groundwater seepage is focussed causing failure along bedding planes and the generation of large deep-seated rotational and translational slides.

Seasonal groundwater infiltration, migration and seepage appear to be one of the major controls on landslide activation. Coastal erosion at the base of the cliffs also adds to the landslide hazard maintaining a state of marginal stability/instability by reducing the lateral constraining pressures that effectively anchor some of the larger slides.